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THE METEOROLOGICAL MAGAZINE

RETIREMENT OF DR R. J. MURGATROYD, O.B.E.

Dr R. J. Murgatroyd, O.B.E., Deputy Chief Scientific Officer in the Dynamical Climatology Branch of the Meteorological Office, retired from the Office on 16 February 1978. His early career was at the Rugby Radio Station of the General Post Office. While there he studied for and obtained a B.Sc. degree in Electrical Engineering. He transferred to the Patent Office as an Assistant Examiner in 1938 but shortly before the outbreak of war he joined the Royal Air Force Volunteer Reserve and was posted to the Meteorological Office Training School as a Pilot Officer. He served as a forecaster at several RAF stations and rose to the rank of Flight Lieutenant but his most significant war-time postings were to Larkhill for a course in sound ranging and later to various Army units where he was given the job of measuring winds at high altitudes by observing the movement of smoke from shells fired to a height of 30 km by a special 'hyper-velocity' gun and also by the less direct method of observing sound waves from distant ground explosions. This work was the start of the long and detailed studies which have given Dr Murgatroyd a unique knowledge of, and world-wide authority in, the affairs of the upper atmosphere.

His formal transfer to the Meteorological Office as a Senior Scientific Officer in 1946 was followed by postings to Larkhill and then, for three years, to Germany. In 1951, on promotion to Principal Scientific Officer, he took charge of the Meteorological Research Flight (MRF) at Farnborough and saw it through a time of rapid development when new aircraft—the Hastings, Canberra and Varsity—were coming into service and new instruments to exploit their capabilities were being developed. Dr Murgatroyd's varied background made him an ideal leader over the whole spectrum of MRF activities, but he also found time to extend his own studies of the atmosphere far above aircraft heights. By the systematic use of meteorological theory he was able to weld together diverse but scanty experimental data—from balloons, a few rockets, sound propagation, high-level clouds, meteor trails and so on—to produce by 1957 a coherent picture of the distribution of winds and temperatures up to 100 km which has been remarkably little changed by the vastly greater numbers of observations which have accumulated since. This work led to the award of a Ph.D. degree and the L. G. Groves Memorial Prize for Meteorology in 1958.

In 1957 Dr Murgatroyd was promoted to Senior Principal Scientific Officer in the new post of Chief Meteorological Officer, MRF and then in 1962 the value of his research was recognized by his transfer to Special Merit status in the General Circulation Branch, which freed him from administrative duties. The quality of his work at MRF, which had already brought him the Darton Prize of the Royal Meteorological Society in 1956, was honoured by the award of an O.B.E. in 1963. He was not left to pursue his research undisturbed for as his reputation grew there were increasing demands on him as a lecturer, as a member of committees of the World Meteorological Organization, the International Association of Meteorology and Atmospheric Physics and other organizations and as Editor of the *Quarterly Journal of the Royal Meteorological Society*. In 1971 he gained the rare Special Merit promotion to Deputy Chief Scientific Officer and a year later he was appointed Chairman of the Committee on Meteorological Effects of Stratospheric Aircraft (COMESA). This committee was set up to direct a widespread and urgently pursued program to assess the effects of aircraft operations (in particular, Concorde) on the composition and meteorology of the atmosphere. Under Dr Murgatroyd's leadership COMESA brought together work from the universities, government departments and industry that covered all facets of the problem and produced a report of unassailable authority which proved an effective counter to earlier alarming predictions based on inadequate or improperly understood evidence. This work, for which he received the L. G. Groves Prize again in 1975, has proved a fruitful starting point for further studies of the stratosphere, studies which are fortunately not likely to end with Dr Murgatroyd's formal retirement.

These bare facts of an impressive career give little idea of the influence that Dr Murgatroyd has always exerted on those round him. His example of hard and thorough work, looking at every aspect of a problem, but with a sound scientific instinct for those aspects likely to be important, has been all the more effective for the quietness and complete lack of ostentation with which he has carried his considerable reputation and authority. We have all admired his combination of high scientific achievement and unselfish co-operativeness; he has been one of those who set the standards on which the life and work of the Office rest. We wish him and his wife an active and happy retirement.

K. H. STEWART

ACOUSTIC SOUNDING OF RADIATION FOG

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SUMMARY

Acoustic sounder echo patterns typical of those obtained at Cardington during instances of radiation fog are described and discussed. The height of a strong elevated layer echo is shown to be closely correlated with the fog depth. Acoustic estimates of the fog depth are compared with those from the traditional (profile) method and the potential usefulness of this technique for monitoring radiation fog evolution is illustrated by a case study.

1. INTRODUCTION

An important parameter in the forecasting schemes currently used for the prediction of radiation fog clearance is the fog depth, D. The estimated fog clearance temperature, T_2, and dawn temperature, T_1, along with the fog depth enable an estimate to be made of the time required (following dawn) for the fog to clear (Barthram, 1964). The fog depth is usually assessed by inspection of Cardington Balthum temperature and humidity profiles through the fog (Painter, 1970), or, though with much less accuracy, by the use of a represent-ative radiosonde ascent. The potential temperature profiles usually reveal the existence of a strong inversion associated with the fog top. This is formed by the lifting of the nocturnal inversion following fog formation and the continued radiational cooling from the fog top. Above the radiatively shielded ground a superadiabatic lapse rate develops and helps to establish a convective regime within the fog in which soil heat flux is transferred upwards (Roach *et alii*, 1976).

One technique used to determine the height of the fog top at dawn uses information from the midnight Balthum ascent. From the profiles of tempera-ture and dew-point plotted on a tephigram the nose of the temperature inver-sion is raised by 5 mb and the temperature decreased by 1·5 K; this point is then joined to the night minimum temperature by a straight line. The intersection of this line and the dew-point curve provides an estimate of the fog depth at dawn (Heffer, 1965). Clearly this method relies upon conditions being well behaved between the time of the ascent and dawn. Furthermore the accuracy of any prediction deduced from a single ascent depends on the spatial representativeness of the ascent itself. Aircraft and tower observations indicate that the fog top is often perturbed by low frequency waves propagating in the stable air aloft with the result that the surface of the fog takes on a somewhat 'corrugated' appear-ance. Obviously the information obtained about the fog top depends upon the point of traverse through this surface. It is the purpose of this paper to demon-strate that a monostatic acoustic sounder may be usefully employed in the study of radiation fog and also provides an alternative method capable of continually monitoring the fog depth.

The propagation of sound in the atmosphere has been a topic of research for at least one hundred years (see Tyndall, 1874). In 1946 Gilman, Coxhead and Willes of the Bell Telephone Laboratories reported the detection of acoustic echoes of unexpectedly high intensity, much greater than could be accounted for by reflection alone. However, many years elapsed before it was shown by McAllister (1968) that these echoes could be easily obtained and displayed (for ease of appraisal) in height/time form on a facsimile chart recorder. He also

provided some evidence that the echoes were generated by the scattering of sound from inhomogeneities in the temperature field on a scale of about half the wavelength of the transmitted sound.

2. THEORETICAL BACKGROUND

The theory of the scatter of sound by turbulent velocity fluctuations in the atmosphere or by fluctuations in scalar atmospheric variables such as temperature and humidity has been investigated by several authors, for example Lighthill (1953), Kraichnan (1953), Kallistratova (1959) and Monin (1962). Following Monin, the scatter of sound in dry air by inhomogeneities can be expressed, assuming a Kolmogorov spectrum of turbulence, by

$$\sigma_{\rm T}(\theta) = 0.03 \ k^{1/3} \ \cos^2 \theta \left[\frac{C_V^2}{C^2} \cos^2(\tfrac{1}{2}\theta) + 0.13 \ \frac{C_{\rm T}^2}{T^2} \right] (\sin \tfrac{1}{2}\theta)^{-11/3} \qquad .. \quad (1)$$

where $\sigma_{\rm T}(\theta)$ is the scattered power per unit volume, per unit incident flux, per unit solid angle at a direction θ from the initial propagation direction; k is the wave number of the acoustic transmitted wave; C and T are the mean velocity of sound and mean temperature in the scattering volume respectively; C_V^2, $C_{\rm T}^2$ are the structure parameters for the wind and temperature fields, defined by the expressions

$$C_V^2 = \overline{[V(x) - V(x + r)]^2}/r^{2/3}$$

$$C_{\rm T}^2 = \overline{[T(x) - T(x + r)]^2}/r^{2/3} \ .. \qquad .. \qquad .. \qquad .. \qquad .. \qquad .. \quad (2)$$

where $V(x)$ is the instantaneous wind speed at a point x in the direction x to $x + r$, $T(x)$ is the instantaneous temperature at point x, and $V(x + r)$ and $T(x + r)$ are the corresponding instantaneous values at point $x + r$. For backscattered sound (i.e. a monostatic sounder configuration with co-located transmitter and receiver) $\theta = 180°$ so reducing equation (1) to

$$\sigma_{\rm T}(180) = 0.008 \frac{C_{\rm T}^2}{T^2} \lambda^{-1/3}, \qquad .. \qquad .. \qquad .. \qquad .. \qquad .. \qquad .. \quad (3)$$

thereby implying that the echo strength is determined, in this case, by the temperature structure parameter $C_{\rm T}^2$ while also being weakly dependent on the wavelength of the transmitted sound, λ. Equation (3) has been shown to be a fairly good approximation by recent experimental work; see for example Asimakopoulos *et alii*, 1975 and 1976; Neff, 1975; and Haugen *et alii*, 1975. The results have indicated that the intensity of the backscattered sound may be used to obtain a reasonable estimate of $C_{\rm T}^2$. Caughey *et alii* (1978) have also shown that in convective conditions satisfactory estimates of C_V^2 can be obtained from bistatic acoustic sounder returns by using equation (1). The theory outlined above ignores the contributions to the scattered sound from humidity fluctuations and the reflection of sound from sharp gradients of refractive index. However, it appears that in many cases these terms are typically one or two orders of magnitude lower than the temperature and velocity contributions and can therefore be ignored (Little, 1969). In fog the possibility exists of an additional contribution to the echo intensity due to Rayleigh scattering from the fog

water droplets. Using an approximate form of Rayleigh's (1872) theory for the scatter of sound by a spherical particle whose diameter (d) is much smaller than the acoustic wavelength (λ) (a good approximation for fog since $d/\lambda \approx 10^{-5}$), Little (1972) has shown that the scattering cross-section per unit volume is

$$\sigma_D(\theta) = \frac{\pi^5}{9\lambda^4}\, [1 - \frac{3}{2}\cos\theta]^2 \left(\frac{1}{V}\sum_V d^6\right), \quad .. \qquad .. \qquad .. \qquad .. \qquad .. \qquad (4)$$

and hence for backscattered sound

$$\sigma_D(180) = \frac{25\pi^5}{36\lambda^4}\left(\frac{1}{V}\sum_V d^6\right). \qquad .. \qquad .. \qquad .. \qquad .. \qquad .. \qquad .. \qquad (5)$$

For the scatter from fog droplets to be detectable it must exceed the noise limit of the system. This is determined principally by the local environmental noise level (i.e. wind noise, traffic noise, etc.) rather than the limiting thermal noise from the random motion of atmospheric molecules or the electron shot noise in the receiver electronics. The quantitative comparisons at Cardington between acoustic echo intensities and the direct measurement of C_T^2 have shown that this environmental noise generally limits the smallest C_T^2 values measurable to something greater than 10^{-6} K^2 m$^{-2/3}$. From equation (3), then, the minimum scattering coefficient detectable with the present Cardington sounder is of the order of 10^{-13} m^{-1}. Measurements available of fog droplet spectra (Brown, private communication) permit an estimate of the typical scattering cross-section $\sigma_D(180)$ to be expected from radiation fogs. The value obtained of 10^{-15} m^{-1} is two orders of magnitude smaller than the minimum detectable and hence it would seem that the echoes recorded by the Cardington monostatic sounder in fog conditions are generated principally by small-scale temperature inhomogeneities. It is worth noting, however, that for scattering through ninety degrees $\sigma_T(90)$ is zero whereas $\sigma_D(90)$ is still appreciable and thus the possibility remains that an acoustic system could be constructed which would monitor only the presence of water droplets or other hydrometeors.

3. Experimental details

The Cardington Acoustic Sounder has been described in a previous article (Crease *et alii*, 1977) so only a brief outline is given here. The acoustic array consists of a 6×6 arrangement of small, re-entrant horn loudspeakers separated by the half wavelength of the transmitted sound, which, with an operating frequency in the range 1·5–2·0 kHz, is about 0·1 m. Short-duration (about 30 ms), high-intensity bursts of sound are directed vertically into the atmosphere at fixed intervals of typically 2·0 seconds. After loudspeaker ringing has ceased, the array is switched to the listening mode and any echoes received are amplified (by a factor of up to 10⁷) and displayed on a facsimile recorder.

During the 1976/77 winter the acoustic sounder was put into operation whenever conditions appeared favourable for the formation of radiation fog. In addition Balthum ascents were made at regular two-hourly intervals to obtain profiles of wind speed, temperature, humidity and pressure across the depth of interest. Echo patterns typical of those obtained in radiation fog show a strong layer echo overlying a region usually exhibiting evidence of convective activity (Crease *et alii*, 1977). This supports evidence from direct measurements of turbulent mixing within fogs (Roach *et alii*, 1976). The layer echo is considered

to be generated by the interaction of convective plumes with the sharp tempera-
ture gradients associated with the fog top and by temperature fluctuations
resulting from breaking wave activity. Clearly radiative cooling and the region
of temperature inversion will extend some distance into the fog and so the exact
relationship between the position of the layer echo and the fog top is difficult
to specify.

4. COMPARISON OF SOUNDER AND PROFILE ESTIMATES OF FOG DEPTH

Over the period October 1976 to February 1977 a total of four cases of radiation
fog were studied. The fog depth was taken as the lower bound of the layer echo
in the sounder returns. From the profiles of temperature and humidity the
highest level with zero wet-bulb depression and the first level to indicate signifi-
cant temperature increase and wet-bulb depression were averaged for an estimate
of the fog top. This latter method introduces a tolerance dependent on the
vertical resolution of the Balthum profile. In the present case this implies an
error of about ± 60 m on the fog depth estimate. The tolerance on the sounder
estimate depends upon the width of the acoustic pulse, which will act to broaden
a narrow region of intense thermal activity, as well as the oscillations in the
height of the layer echo which occur on time scales varying from a few minutes
to several hours. The comparison of fog depth estimates for all available
occasions is given in Figure 1, the typical tolerances on the estimates being
denoted by the error bars. Generally good agreement is indicated for fog depths
between 60 metres and 240 metres. On one occasion when a droplet spectro-
meter was available the actual height of the fog top was monitored over a three
hour period on the early morning of 27 October 1976. The results (see Figure 2)
demonstrate a clear correlation between the height of the layer echo and that
of the actual fog top.

5. CASE STUDY OF RADIATION FOG ON 14/15 NOVEMBER 1976

To illustrate the potential of acoustic sounding in monitoring the evolution of
radiation fog a description is given of the sequence of events which occurred
between fog formation and dissipation on 14/15 November 1976.

(a) *Synoptic situation*

On 14 November 1976 a ridge of high pressure extended south-westwards
across south-east England from an anticyclone centred over Scandinavia. In
the East Midlands and East Anglia fog which had formed in the early morning
persisted throughout the day in many districts, whereas in the Cardington area
it had cleared by midday. With light winds, moist air and nocturnal cooling
conditions were conducive to the reformation of radiation fog. By 1550 GMT
ground fog had formed and this then began to deepen and thicken. During the
night a weak frontal system moved into south-west England and Wales and
associated patches of stratocumulus cloud were evident in the south Bedford-
shire area from about midnight onwards.

(b) *Relationship between acoustic sounder facsimile records, wind and temper-
ature profiles and surface net radiation*

The sequence of low-level wind speed and temperature profiles during the early
period of fog formation is shown in Figure 3. As the fog began to deepen after
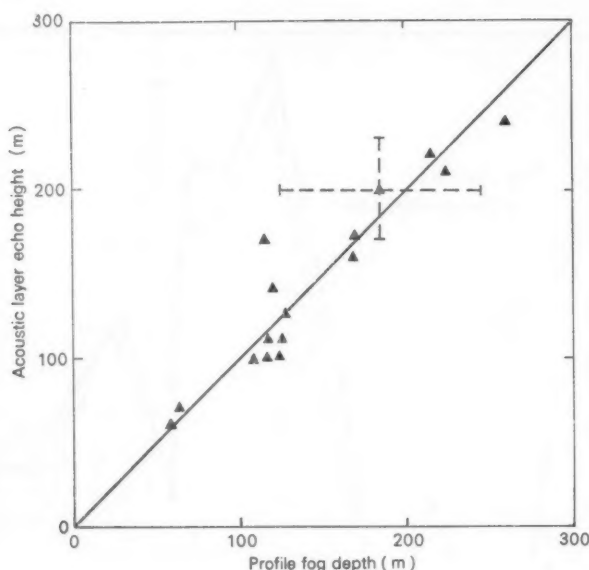
FIGURE 1—COMPARISON OF ACOUSTIC SOUNDER LAYER ECHO-HEIGHT AND ESTIMATES OF FOG DEPTH FROM TEMPERATURE AND HUMIDITY PROFILES

1800 GMT the surface-based radiation inversion started to dissipate owing to the gradual decrease in long-wave radiational cooling and the soil heat flux acting to raise the surface temperature. During this period the 0·5 m temperature rose by ≈ 2 °C whereas at 16 m the temperature fell. Mean winds up to 8 m height remained in the range $\frac{1}{2} - 1\frac{1}{2}$ m s^{-1}. The sharply rising echo layer on the acoustic facsimile chart (Figure 4) evident at around 1830 GMT is most probably associated with the fog top and it is clear that when the fog depth exceeded about 100 m the Richardson number in the surface layer had fallen to $\approx 0·1$ (eventually becoming negative around 2200 GMT) and the net radiation increased to near zero.

Shown in Figure 5 is the sequence of Balthum soundings obtained on this night. The early evening profiles show the presence of a strong nocturnal inversion extending up to about 30 m. By 2038 GMT, however, with the formation and deepening of the fog, the profiles had altered significantly and indicated the presence of a fairly deep saturated adiabatic layer. On the next ascent (2310 GMT) the depth of this layer is maintained but in subsequent profiles some decrease is apparent so that by 0626 GMT a surface-based inversion was again in evidence and the fog had dispersed.
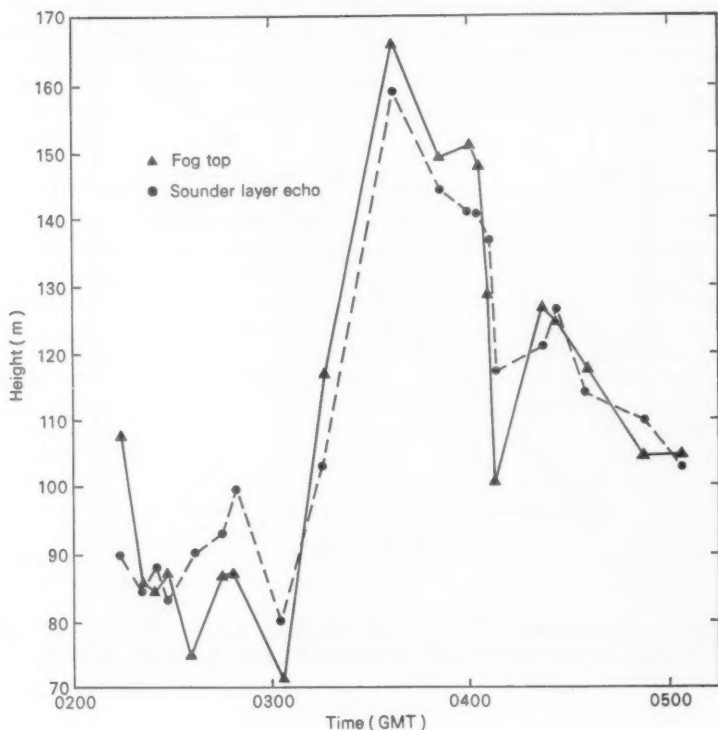
FIGURE 2—COMPARISON OF ACOUSTIC SOUNDER LAYER ECHO-HEIGHT (●) AND MEASUREMENTS OF THE FOG TOP HEIGHT (▲), OBTAINED BY USING A BALLOON-BORNE DROPLET SPECTROMETER, BETWEEN 0200 AND 0500 GMT ON 27 OCTOBER 1976

These implied variations in the fog depth correlate well with the movement of the layer echo height on the sounder facsimile chart (see Figures 4 and 6). This shows a gradually deepening fog up to about 2100 GMT after which time a slow and erratic decrease begins with the layer echo becoming generally more tenuous and broken. A gradual improvement in visibility began at around 0500 GMT reaching 500 m by 0600 GMT and 1000 m shortly afterwards, when a nearly complete layer of stratocumulus cloud at around 1700 m was apparent. Consistently with these developments the acoustic layer echo became progressively weaker and more diffuse, gradually merging with other echoes as the general echo pattern returned to one more typical of a stable boundary layer.

6. CONCLUDING REMARKS

This paper has demonstrated that an acoustic sounder may be usefully employed in the study of radiation fog and in particular enables reasonably accurate
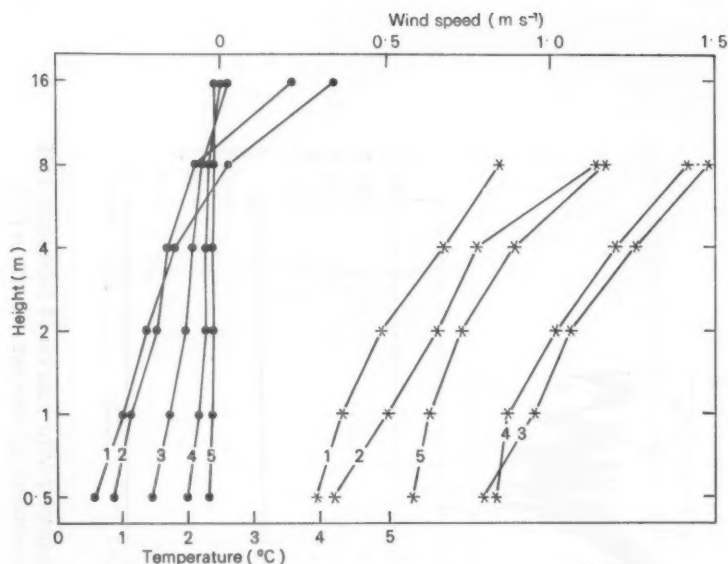
FIGURE 3—PROFILES OF WIND SPEED (∗) AND TEMPERATURE (●) DURING THE
INITIAL PERIOD OF FOG FORMATION ON 14 NOVEMBER 1976

The profiles are consecutive ten minute averages over the period 1750 to 1840 GMT.

estimates of the depths of established fogs to be made. Additionally since the
sounder can be used continually it provides valuable information on the evolu-
tion of the fog and in the period following dawn the weakening of the layer
echo can be monitored and the progress towards fog clearance observed. A
detailed investigation of the microphysics at the fog top is required in order to
elucidate the relationship between the acoustic layer echo, the temperature
profile, the turbulence field and the actual fog top.

ACKNOWLEDGEMENTS

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COLE, R. S., CAUGHEY, S. J.　　　　　returns and the direct measurement of atmospheric
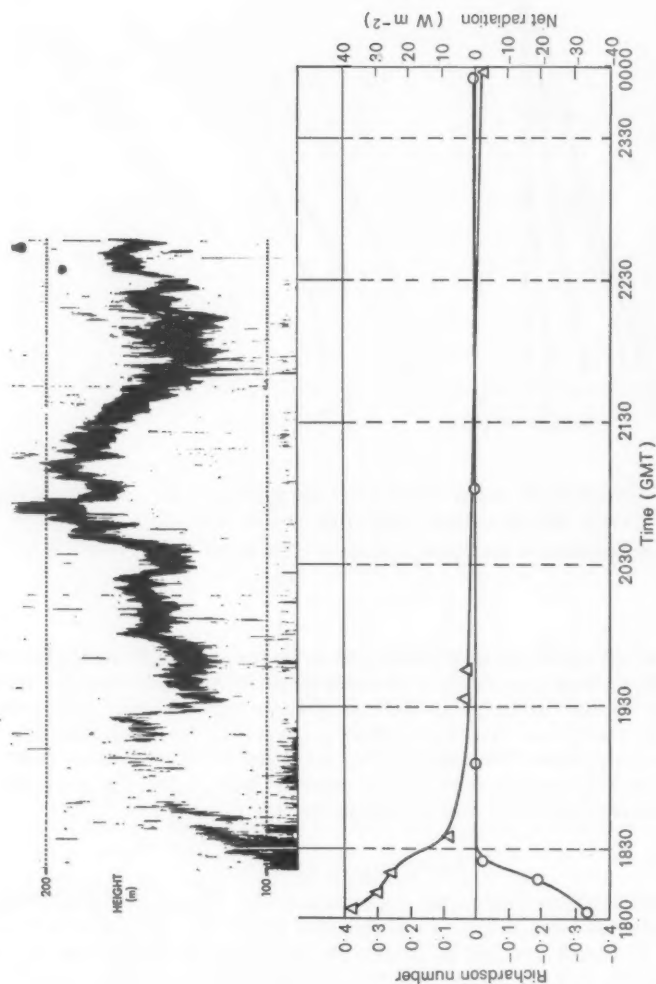and CREASE, B.A.　　　　　　　　　temperature fluctuations. *Boundary Layer Meteor-
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FIGURE 4—VARIATION OF ACOUSTIC SOUNDER LAYER ECHO-HEIGHT WITH TIME ON 14 NOVEMBER 1976 AND COMPARISON WITH SURFACE LAYER RICHARDSON NUMBER ($\triangle$) AND NET RADIATION ($\bigcirc$)
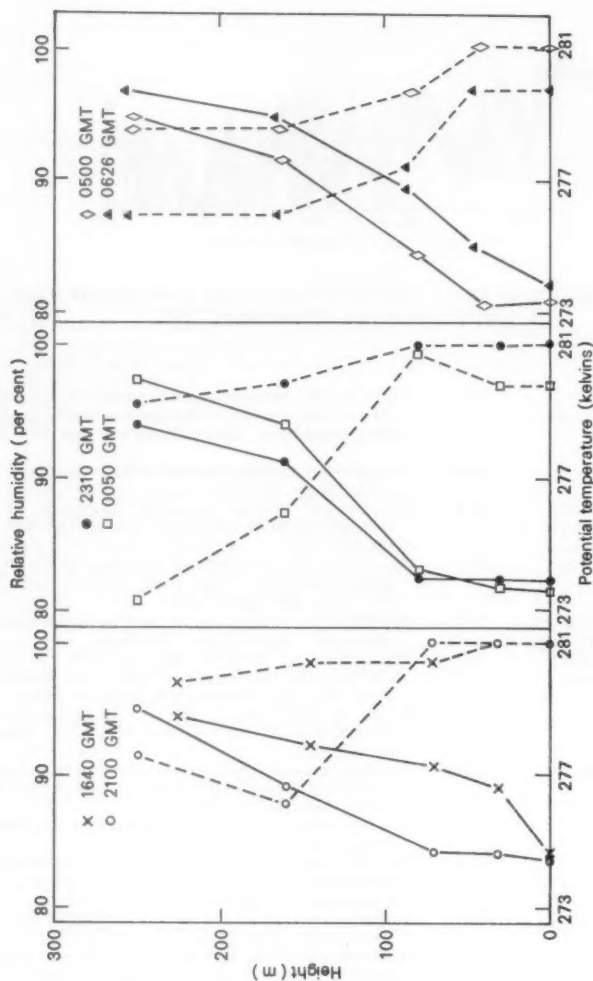
FIGURE 5—SEQUENCE OF BALTHUM PROFILES OF TEMPERATURE AND HUMIDITY FOR 14/15 NOVEMBER 1976
—— Potential temperature – – – Relative humidity
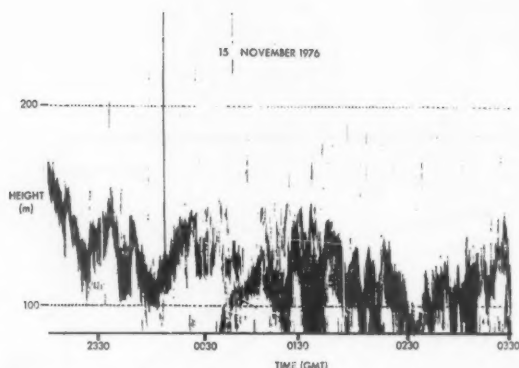The time of the second profile should read 2038 GMT and not 2100 GMT.

FIGURE 6—VARIATION OF THE ACOUSTIC SOUNDER LAYER ECHO-HEIGHT WITH TIME
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THE FORECASTING OF OROGRAPHICALLY ENHANCED RAINFALL ACCUMULATIONS USING 10-LEVEL MODEL DATA

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SUMMARY

A diagnostic model has been developed in order to estimate the effect of orography on surface rainfall accumulations. The model calculates rainfall amounts at grid points $3\frac{1}{4}$ km apart using large-scale input parameters forecast by the fine-mesh version of the Meteorological Office 10-level model. The performance of the orographic model during a two-week trial period is assessed.

1. INTRODUCTION

A major problem in the provision of short-period weather forecasts is the estimation of local rainfall which may be modified by local topographic effects. An important operational requirement is the forecasting of rainfall accumulations in hilly areas (Holgate, 1973). Methods of forecasting based on the extrapolation of the movement of precipitation echoes derived from radar observations (Hill, Whyte and Browning, 1977) are only useful for forecasting up to 6 hours ahead, and even these forecasts will require modification if the local orographic effects are large. For forecasts up to 36 hours ahead a numerical model is required. Unfortunately the resolution of present numerical models is one or two orders of magnitude greater than the scale of the orographic features, and the effect of these features is not incorporated into the resulting rainfall forecasts. One solution is to use the output from a numerical model as input to a separate, numerically simple model that has a grid size small enough to resolve the orography adequately and which contains the essential physics of the orographic rainfall process.

Collier (1975) has formulated a model of orographic rainfall which provides for an elaborate method of estimating vertical velocity in the vicinity of orographic features. This model has been employed with some success to estimate rainfall accumulations in North Wales, radiosonde observations having been used as input data. As Collier (1977) indicates, the model results are only valid in special circumstances when there is efficient conversion of orographic cloud to liquid rain water. The cloud microphysical processes which are not parametrized in Collier's model play an important role in determining the extent of the orographic influence on the rainfall. The model about to be described is similar in concept to Collier's but a simpler formulation for obtaining local vertical velocity is used together with a scheme for treating the cloud physics processes. It is based on a scheme suggested by Jonas (1976). The method

described by Bader and Roach (1977) for calculating the washout of droplets by raindrops from a higher cloud in a low-level orographically produced cloud has been adapted for use in this model.

2. MODEL FEATURES

(a) *General description*

The model has been designed to use the output from the fine-mesh version of the Meteorological Office 10-level model (Burridge and Gadd, 1977); although the latter includes orography, it is in a very smooth form. Local perturbations of the flow due to sub-gridscale variations in orography are assumed not to affect the large-scale flow, but to add an extra component to the vertical velocity and modify the relative humidity at the fixed pressure levels.

Precipitation is formed in a given layer by adiabatic ascent. It is assumed that the rate of rainfall so formed may be written as

$$P_1 = -k_1 k_2 \, \omega (\partial r_s / \partial p)_{sat} \, \rho \Delta z \text{ g s}^{-1} \text{ m}^{-2}. \quad .. \quad\quad .. \quad\quad .. \quad\quad .. \quad\quad (1)$$

This is similar to the expression derived by Collier (1975); $\omega = \mathrm{d}p/\mathrm{d}t$ is the vertical velocity in p co-ordinates, $(\partial r_s / \partial p)_{sat}$ is the derivative of saturated humidity mixing ratio with respect to pressure for a saturated adiabatic ascent, ρ is the air density and Δz the thickness of the layer or, in the case of the lowest level, the height of the level above the surface. The derivation of the parameters k_1 and k_2, which depend on the way in which the relative humidity is modified by ascent, will be described in section 2(c).

Precipitation falling into the next layer is enhanced by the accretion of cloud water. This washout process increases the rate by P_2 to give a total rate of rainfall $P = P_1 + P_2$, where we assume that $P_2{}^j$ is a function of P^{j-1}, the precipitation rate in the layer ($j-1$) above, and of q^j, the cloud liquid water content in the layer j. Details of this process will be considered later in section 2(d). The surface rainfall is calculated by summing the contributions from five 100 mb layers, from the surface to 500 mb.

The large-scale velocity components (u, v, ω), the layer thickness Δz, and the humidity mixing ratio r, available from the 10-level model are interpolated to the orographic model grid points whose resolution is $3\frac{1}{3}$ km. The orographic height is defined at each point.

(b) *Formulae for vertical velocity and* $(\partial r_s / \partial p)_{sat}$

A simple parametrization of the effect of orography has been adopted. The vertical velocity ω is given by

$$\omega = \omega_L + \omega_T \text{ mb s}^{-1}, \quad\quad .. \quad\quad .. \quad\quad .. \quad\quad .. \quad\quad .. \quad\quad .. \quad\quad (2)$$

where ω_L is the large-scale vertical velocity and ω_T is an orographically induced component assumed to be proportional to the scalar product of horizontal velocity V and local topographic gradient ∇H, that is

$$\omega_T = -k_3 \, (V \cdot \nabla H) \, \rho g \text{ mb s}^{-1}. \quad .. \quad\quad .. \quad\quad .. \quad\quad .. \quad\quad .. \quad\quad .. \quad\quad (3)$$

The effect of orography is assumed to reduce linearly to zero at 500 mb. Accordingly the factor k_3 varies linearly with the height of the level above the surface, from zero at 500 mb, to unity at the surface. This treatment is very crude but seems justified as a first attempt because the overall accuracy of the forecast is dominated by the accuracy of the 10-level model results. The effect

of stability on the magnitude and vertical extent of the ascent is also neglected.

The rate of change of saturated mixing ratio with respect to pressure along a saturated adiabat may be readily derived from thermodynamic arguments. The energy conservation equation may be written as

$$L\,dr_s = c_p dT + (RT/p)dp, \qquad \ldots \qquad \ldots \qquad \ldots \qquad \ldots \qquad \ldots \qquad (4)$$

where L is the latent heat of vaporization and c_p the specific heat at constant pressure.

In terms of the saturated vapour pressure e_s, the humidity mixing ratio is $r_s \approx \epsilon e_s/p$, where $\epsilon = 0 \cdot 622$. In differential form this becomes

$$dr_s/r_s = de_s/e_s - dp/p. \qquad \ldots \qquad \ldots \qquad \ldots \qquad \ldots \qquad \ldots \qquad (5)$$

The term e_s may be eliminated from equation (5) by using the Clausius–Clapeyron relation to give

$$dr_s/r_s = \epsilon L dT/RT^2 - dp/p. \qquad \ldots \qquad \ldots \qquad \ldots \qquad \ldots \qquad \ldots \qquad (6)$$

Manipulation of equations (4) and (6) gives

$$(\partial r_s/\partial p)_{sat} = r_s (RT/p)(\epsilon L - c_p T)/(L^2 \epsilon r_s + Rc_p T^2). \qquad \ldots \qquad \ldots \qquad \ldots \qquad (7)$$

(c) *Modification of the relative humidity by forced adiabatic ascent.*

The orographically induced displacement is parametrized in the same way as the orographically induced component of vertical velocity. At a given level, the displacement is given by $H_T = k_3 H$, where k_3 is as in section 2(b) and H is the orographic height.

The relationship between the final relative humidity X_f and the initial relative humidity X_i for a dry adiabatic ascent H_T is appoximately linear, i.e.

$$X_f = X_i \{1 + \alpha(p,T) H_T\}, \qquad \ldots \qquad \ldots \qquad \ldots \qquad \ldots \qquad \ldots \qquad (8)$$

where α is a function of temperature and pressure only.

The parameter k_1 (see equation (1)) is assumed to depend on the vertical velocity and the relative humidity. Thus $k_1 = 0$ if $\omega > 0$, that is where there is local descent or if $X_f < 1$ when the air is unsaturated after ascent. Otherwise k_1 is set equal to unity and precipitation is allowed to form.

The parameter k_2 is introduced in an attempt to take into account the depth of the cloud. If the air is already saturated before orographic uplift then $k_2 = 1$. The parameter k_2 is dependent on the length of time the growing droplets spend in the orographic cloud. The vertical displacement H_s to reach saturation is given by substituting $X_f = 1$ in equation (8)

$$H_s = (1 - X_i)/\alpha X_i. \qquad \ldots \qquad \ldots \qquad \ldots \qquad \ldots \qquad \ldots \qquad (9)$$

The droplets are assumed to follow the air motion with a vertical velocity $-\omega/\rho g$ m s^{-1} for the remainder of the forced displacement $(H_T - H_s)$. Hence the time taken for the droplets to grow in the cloud is given by

$$t = -\rho g/\omega \{H_T - (1 - X_i)/\alpha X_i\}. \qquad \ldots \qquad \ldots \qquad \ldots \qquad \ldots \qquad (10)$$

A typical time scale for droplet growth is 20 minutes, thus we assume

$t > 1200$ seconds	$k_2 = 1$
$300 \leqslant t \leqslant 1200$ seconds	$k_2 = (t-300)/900$
$t < 300$ seconds	$k_2 = 0.$

If air becomes saturated during ascent the remainder of the ascent is taken along a saturated adiabat. The final value of saturated mixing ratio r_s is

derived from equation (7). Hence a new value of the relative humidity $X_t = (r/r_s)$ is found which is greater than 100%. It is used to derive an estimate of the cloud liquid water content q, assuming (i) that 10% of the condensed water is retained in the cloud and (ii) that cloud forms at 90% relative humidity.

$$q = 0\cdot1\ (X_t - 0\cdot9)r_s \times 10^{-3}\ \rho\ \text{kg m}^{-3}. \quad .. \qquad .. \qquad .. \qquad .. \qquad .. \quad (11)$$

(d) *Washout*

Now that an estimate of cloud liquid water has been made the rainfall P_2 due to accretion of cloud liquid water may be calculated. For simplicity a single cloud drop size of radius 10 μm is assumed. For a raindrop of radius a, with a fallspeed V_a, and collection efficiency E_a, the rate of accretion W' of cloud liquid water is simply $W' = E_a V_a \pi a^2 q$, since $\pi a^2 V_a$ is the volume of air swept per second. If the raindrop distribution is $N_a \Delta a$, then the total washout summed over all drops is given by

$$W = q \sum_a N_a V_a E_a \pi a^2 \Delta a. \qquad .. \qquad .. \qquad .. \qquad .. \qquad .. \qquad .. \quad (12)$$

This expression is used by Bader and Roach in their model. $N_a \Delta a$ is defined by the 'Best' dropsize distribution which is a function only of precipitation rate (see Mason 1971). The terminal velocities and efficiency factors are also taken from Mason. The summation in equation (12) is then only a function of precipitation rate. The increase in precipitation rate is $P_2 = W \Delta z$, where Δz is the layer thickness.

Thus the total precipitation rate derived within a layer is

$$P = P_1 + P_2 = -k_1 k_2 \omega (\partial r_s / \partial p)_{\text{sat}}\ \rho \Delta z + W \Delta z\ \text{g s}^{-1}\ \text{m}^{-2}. \quad .. \qquad .. \quad (13)$$

This is subject to the constraint that the maximum allowable rainfall rate is $P = -\omega (\partial r_s / \partial p)_{\text{sat}}\ \rho \Delta z$.

(e) *Other features of the model*

Three other features included are evaporation, precipitation drift and spatial averaging.

If rain falls through a layer of unsaturated air, some or all of it will evaporate. The evaporation scheme uses empirically derived relationships due to Best (1952), a version of which is also used in the 10-level model. A raindrop of radius a_1 reduces to a radius a_2 when falling from height z_1 to z_2.

The empirical relation between a_1 and a_2 is

$$a_1^2 - a_2^2 = E(z_1, z_2)\ (1 - X)^{1\cdot13}, \quad .. \qquad .. \qquad .. \qquad .. \qquad .. \qquad .. \quad (14)$$

where X is the relative humidity and E is an empirical parameter derived by Best which depends on the height of the levels. The initial precipitation rate is resolved into a dropsize spectrum as in section 2(d), a new dropsize spectrum is computed using equation (14) and from this the final precipitation rate can be derived.

The horizontal drift of precipitation as it falls can have a considerable effect on the distribution of surface rainfall. Although the thermodynamics of the ice phase is not included in the model, the precipitation is considered as snow for the purposes of calculating drift if the temperature of the layer is less than 273 K. For example, the time taken for snow to fall from one level to the next is about 1000 seconds assuming a fallspeed of 1 m s^{-1}. In a strong wind of 25 m s^{-1}

the snow will have drifted 25 km (that is to say about 7 grid lengths). If the precipitation is in the form of rain with a fallspeed of approximately 5 m s^{-1}, the drift is correspondingly reduced but is still appreciable.

Finally to avoid unrealistic gradients which might arise since the rainfall rate at each grid point is calculated independently of its neighbours, a 1–2–1 smoothing operator is applied in both the x-direction and the y-direction at each internal point.

3. RESULTS

The aim of this study has been to provide a potential forecasting model. To assess the performance of the model it was tested over an extended period, data from a 10-level rectangle forecast being used as input. The accuracy of the rectangle forecast limits the accuracy that can be expected from the orographic model; an accurate analysis of the large-scale variables is, however, difficult to obtain on an hourly basis. The method of assessment used here shows how the model would perform in practice.

The orographic model was used to provide forecasts of rainfall over Wales and central England. The area covered by the study is shown in Figure 1. The orientation of this area is the same as that of the fine-mesh 10-level model whose grid points are marked by crosses. Figure 1 also shows the orography used in the calculations and the 23 areas of size 2500 km^2 used for verification. The trial was for 14 consecutive days starting on 3 October 1976.

Throughout the period a westerly or south-westerly situation persisted with fronts and depressions crossing the British Isles, notably on the 14th when a deep depression reached south-west England and moved slowly north-east.

The input data for the orographic model were extracted hourly from the 10-level model fine-mesh forecast based on a midnight analysis. The data from $T + 9$ to $T + 32$ hours were used, so that the forecast period of the orographic model coincides with a rainfall day. The large-scale parameters used as input to the model were assumed to be representative of conditions for a 60-minute period and were updated each hour; 24-hour accumulations were obtained by summing the predicted hourly rainfall amounts.

The only suitable observational data with which to compare the model results were the daily rainfall totals from the national rain-gauge network. Programs developed by the hydrometeorological section of the Meteorological Office produce objective estimates of the areal rainfall totals (Shearman, 1975). They may be in error in data-sparse areas but in the absence of any other information they will be considered as 'truth'. A further problem with verification is that the observed rainfall is the sum of convective type and dynamic type. The orographic model only caters for dynamically induced ascent (both large scale and orographically induced), therefore when comparing areal totals a contribution to the total due to convective processes was included.

This convective contribution was derived from the 10-level model deep convection scheme (Hayes, 1977) which added about 20 per cent to the rainfall totals for low-lying areas. This figure is in broad agreement with reports from synoptic stations although it is probably an overestimate for inland stations.

Since the accuracy of the results from the orographic model depends directly on the quality of the forecast input data derived from the 10-level model, verification of the results would be incomplete without a comparison with the results from the present operational scheme.
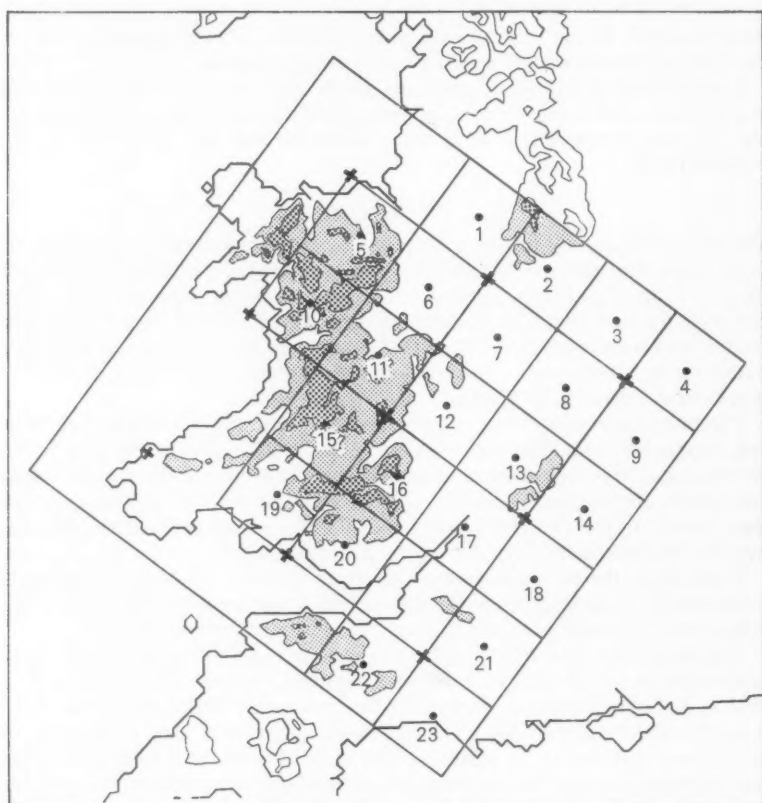
FIGURE 1—OROGRAPHY ON A $3\frac{1}{2}$ km GRID

Light shaded areas represent land between 200 m and 400 m above mean sea level.
Dark shaded areas represent land over 400 m above mean sea level.
X rectangle grid point   ● centre of 50 × 50 km areas used for verification

Figure 2 shows the difference between model and rain-gauge estimates of the
2 week totals for each of the 23 areas of size 2500 km² as a function of gauge
estimate. Figure 2(a) refers to the 10-level model and Figure 2(b) refers to the
orographic model. The areas are numbered as in Figure 1 for reference, lowland
areas being marked with crosses and areas where a significant percentage of the
land is above 200 m being marked with dots. The closer the points to the *x*-axis
the better the forecasts. Several conclusions can be drawn from Figure 2(a).
It is evident that although the 10-level model forecasts rain fairly accurately
for most of the low-lying areas, as the orographic influence becomes more
pronounced it increasingly under-forecasts the rainfall. The improvement given
by using the orographic model is greatest for the hilly areas. For the eight areas
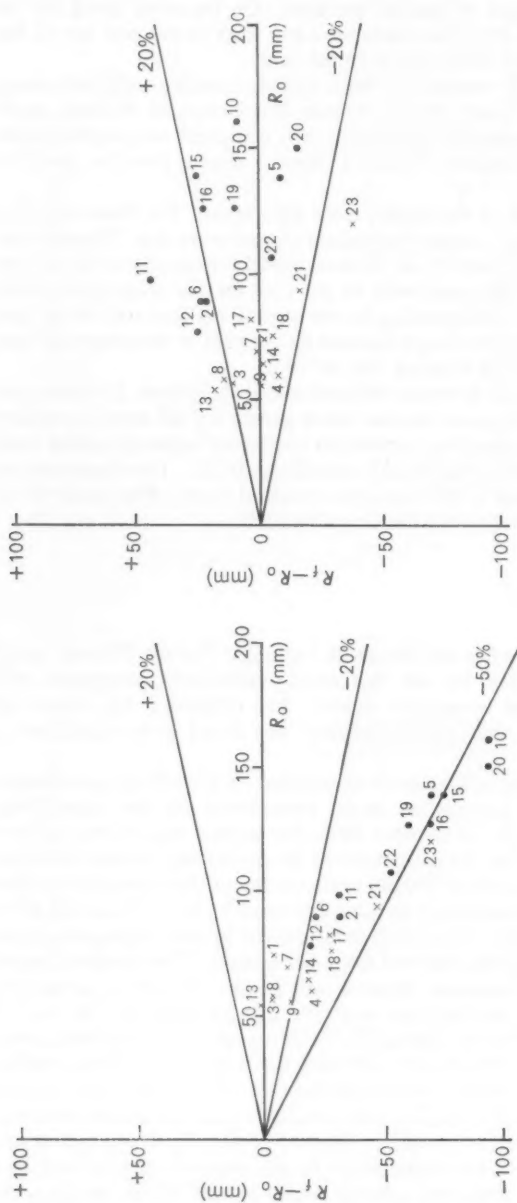
(a) 10-level model

(b) Orographic model

FIGURE 2—PLOT OF (FORECAST RAINFALL MINUS OBSERVED RAINFALL) AGAINST (OBSERVED RAINFALL) FOR TWO WEEK PERIOD

(a) compares results from 10-level model with observed rainfall.
(b) compares results from orographic model with observed rainfall.
● hilly areas X lowland areas
The straight lines show the bounds for forecast errors of 20% and 50% respectively.

where the rainfall exceeded 100 mm in two weeks, the 10-level model under-forecast the rain by between 45 and 60 per cent. On the other hand for the orographic model (Figure 2(b)) the errors were less than 18 per cent for all but one area and five areas had errors below 10 per cent.

Figure 3 shows the daily rainfall for both models together with rain-gauge estimates for area 20 in South Wales. Figure 3(a) compares 10-level model results with rain-gauge estimates and Figure 3(b) compares orographic model results with rain-gauge estimates. Figure 4 shows a similar plot for area 5 in North Wales.

These results are typical of the results from hilly areas. For both areas the orographic model produced a better forecast on almost every day. The per cent-age of observed rainfall forecast by the 10-level model for area 5 was 51 per cent for the two week period compared with 94 per cent for the orographic model forecasts. For area 20 the corresponding figures were 40 per cent and 89 per cent. For all 23 areas together the rectangle forecast 62 per cent of the observed rain-fall and the orographic model forecast 106 per cent.

The correlation coefficient between 10-level model daily area forecasts and observed rainfall was 0·580 over the two week period for all areas combined (322 forecasts). The corresponding correlation coefficient between rainfall fore-cast by the orographic model and actual rainfall was 0·702. This improvement in the correlation coefficient is 2·6 times the standard error. The mean of the absolute difference between forecast and actual rainfall

$$M_D = \frac{1}{14} \sum_1^{14} |(r_m - r_a)|$$

was calculated for both models and for areas 5 and 20. For the 10-level model M_D was 5·85 and 6·53 mm for the two areas respectively, compared with 3·56 and 3·03 mm for the orographic model. The difference was tested for significance using a *t*-test and the improvement was found to be significant at the 1 per cent level.

Figures 5 and 6 show the rainfall fields as depicted by the rain-gauge network and as produced by the orographic model respectively for the rainfall day commencing at 09 GMT on 14 October 1976, the wettest day during the trial period. Bearing in mind that the basic input to the model was forecast data and that those data were on a scale of 100 km so that many of the mesoscale features producing intense precipitation (such as those described by Browning *et alii*, 1974) were not defined, we see that the rainfall field forecast by the orographic model fits encouragingly well with that derived from rain-gauges. The distribution and intensity of rainfall in Snowdonia, Exmoor and Pembrokeshire is particularly good. Note the drift of precipitation over the Conwy valley to the east of Snowdon and also the accurate forecast in Pembrokeshire of an enhancement by a factor of three on fairly low hills indicating the importance of the gradient of the hills as well as their height. Although the rain-shadow effect giving drier areas to the immediate lee of the hills is fairly realistic, there is a general tendency to over-forecast the rainfall in the area to the east of the Welsh mountains and it seems that the removal of additional water by precipitation upwind may be a contributory factor. An improved scheme might be one which modifies the large-scale humidity field in the 10-level model to take into account the local removal of water by the orographic model. This would lead to a feedback
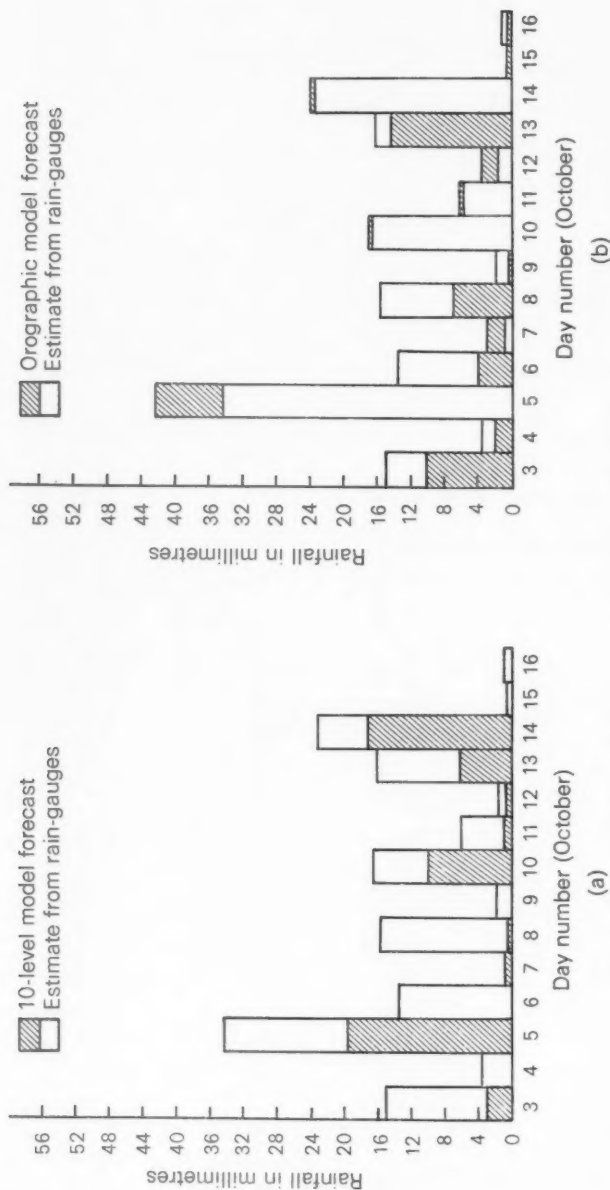
FIGURE 3—COMPARISON OF FORECAST AND OBSERVED RAINFALL IN AREA 20 (SOUTH WALES)

(a) Daily rainfall forecast by 10-level model
(b) Daily rainfall forecast by orographic model
Forecast, and estimated actual, values of rainfall are superimposed with common zero on the axis; the arrangement in the vertical of hatched and blank areas depends on which value is the greater.
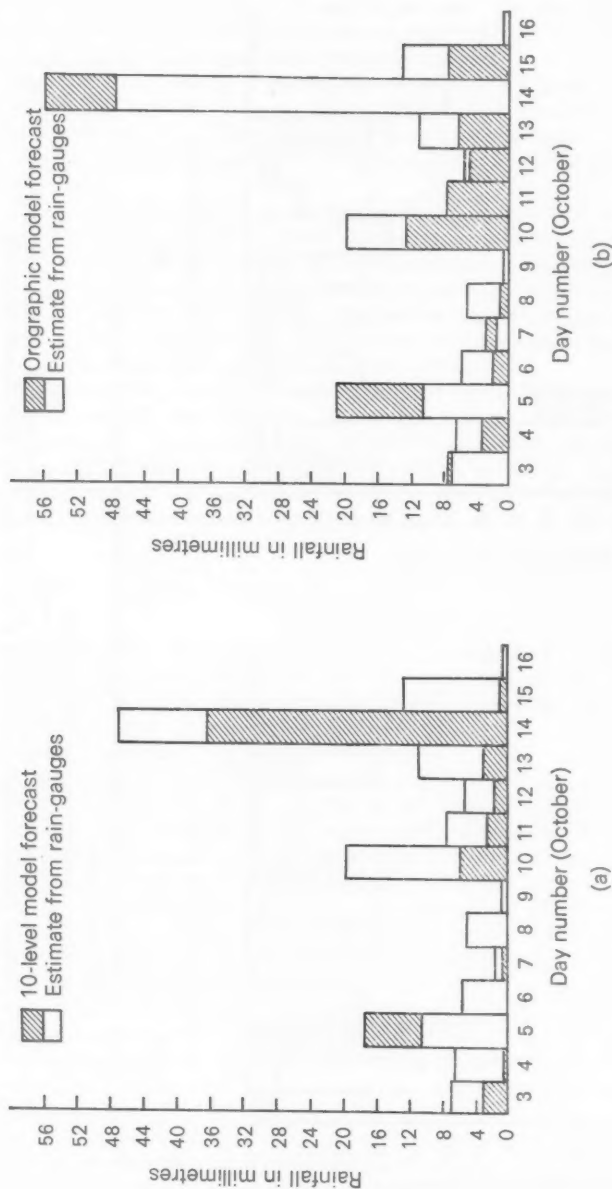
FIGURE 4—COMPARISON OF FORECAST AND OBSERVED RAINFALL IN AREA 5 (NORTH WALES)

(a) Daily rainfall forecast by 10-level model
(b) Daily rainfall forecast by orographic model
Forecast, and estimated actual, values of rainfall are superimposed with common zero on the axis; the arrangement
in the vertical of hatched and blank areas depends on which value is the greater.

FIGURE 5—RAINFALL ESTIMATED FROM RAIN-GAUGES FOR 14 OCTOBER 1976

Horizontal hatching indicates rainfall between 25 and 50 mm, vertical hatching rainfall between 50 and 75 mm, and cross-hatching rainfall in excess of 75 mm.

mechanism between the two models. On 14 October the rain area was forecast by the 10-level model to be too far east and this is reflected in the poor rainfall forecast for the West Midlands (20 mm forecast by both orographic model and 10-level model compared with an observed 5 mm). Since this area is low-lying, the orographic contribution is minimal but a scheme for removing water as just described might have produced better results.

4. CONCLUSION

It has been shown that the orographic influence on rainfall from large-scale systems is well reproduced by this relatively simple model used in conjunction with the 36 hour forecast produced by the fine-mesh version of the 10-level model. Several improvements might be envisaged for the future. The dynamics has been treated very simply and an improved scheme on the lines developed by Collier might be of use. Inclusion of a scheme to handle orographically triggered convection and improved modelling of the rain-shadow effect would also be beneficial.

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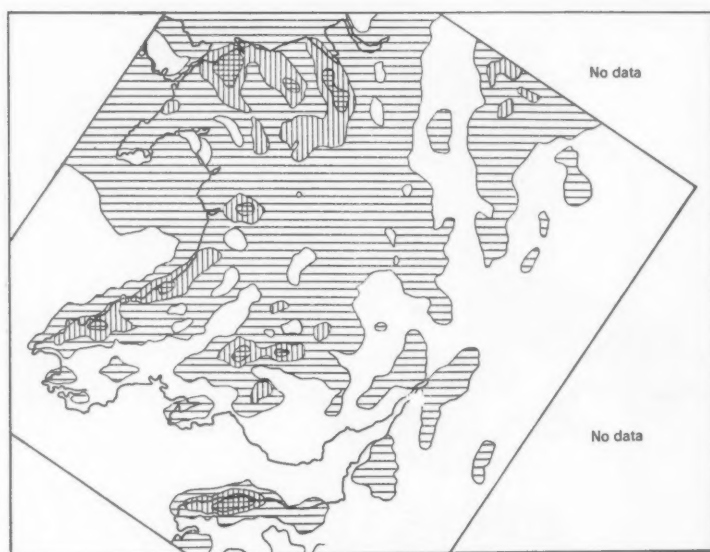|		
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FIGURE 6—RAINFALL FORECAST BY OROGRAPHIC MODEL ON 14 OCTOBER 1976

Horizontal hatching indicates rainfall between 25 and 50 mm, vertical hatching rainfall between 50 and 75 mm, and cross-hatching rainfall in excess of 75 mm.

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A NATIONAL INVENTORY OF WEATHER STATIONS

By S. J. HARRISON

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In the recording of weather and climate the United Kingdom is in the fortunate position of having a central organization for the collating of information. The Meteorological Office receives monthly returns of data from over 600 weather stations. However, it can be argued that even this great number falls well short of an adequate coverage. The problems with a network of this nature are manifold, but given strict standardization of sites, and observation times, there are three which serve to limit the effective value of the data generated:

(1) The stations do not fall into a spatially uniform pattern. Avoidance of difficult terrain such as upland or other inaccessible areas produces a distribution of weather stations which favours lowland locations, particularly those near to coasts. There are, therefore, large gaps in the overall picture of climatic variation in the United Kingdom and with such an unbalanced network, interpolation and extrapolation become necessary tools for the climatologist.

(2) The topography of the United Kingdom is well diversified, small areas containing a complex matrix of localized climates. In hilly areas for example, such as the Pennines, the modifying effects of elevation upon climate can be concealed by those arising from the other elements of topography such as aspect and slope. In such areas, an accurate impression of spatial variation in climate can only be attained from a relatively dense network of weather stations. In reality, the researcher could be considered lucky to find just one station to represent the whole of the area with which he is concerned.

(3) The spatial complexity of climatic variation is exacerbated by the great variability in time of atmospheric conditions over the British Isles. The development of the valley inversion, for example, is obviously dependent not only upon valley configuration but also upon synoptic conditions.

In the light of these problems it would appear that the Meteorological Office network of stations, while it forms the backbone of weather and climate recording in the United Kingdom, contains large gaps which it is desirable to fill. Fortunately this aim can, in some measure, be achieved through the efforts of a number of weather enthusiasts and organizations involved in weather-data collection at both local and national levels.

To some extent, however, the efforts of the weather enthusiast tend to replicate the Meteorological Office network rather than to make significant inroads into some of the larger gaps within it. Organizations such as the Climatological Observers' Link do, however, swell the reservoir of information. Observations of both quantitative and qualitative nature are exchanged between members through the vehicle of monthly bulletins. However, this organization, despite its good works, falls well short of a complete inventory of all weather stations in the United Kingdom.

Contact with research workers in, for example, hydrology, forestry and agriculture, and with teachers' groups has revealed that there is a great deal of hitherto uncollated local weather information in the United Kingdom emanating from stations which often lie in the undesired gaps to which reference has already been made. A major problem with many of these stations is that they frequently do not conform to established standards of siting and instrumentation. Many schools, for example, can only manage a five day schedule of observations. Research organizations, such as the Institute of Hydrology, often rely upon autographic methods of weather recording, occasionally using completely automatic weather stations in upland environments such as on the Cairngorms. However, if full site and instrument particulars are made available, correction can be made for any such deviation from established codes of practice.

Work on an inventory of all weather stations in the United Kingdom began in 1975 with three broad aims. These were firstly to establish the presence of all weather stations not included in Meteorological Office published lists, secondly to determine the siting characteristics of these stations and thirdly to determine the nature of the observations made. The project was named the Register of Weather Stations or ROWS.

To date, information on hitherto unrecorded weather stations has been collected by using a questionnaire method. In addition to particulars of station name, grid reference and height above sea level, the respondent is asked to assess slope, aspect and shelter on five-point scales. He is then asked to indicate whether air temperature, earth temperature, solar radiation, ground surface minimum temperature, precipitation, evaporation, relative humidity, wind speed and wind direction are recorded. The first questionnaires used were four pages long, which elicited useful information on site and instruments but which was generally regarded as being too long. This format has been replaced by one which occupies only one page and which can usually be completed in less than five minutes.

Letters outlining the work of ROWS have so far been published in the *Bulletin of Environmental Education, Area, Geography, Weather*, and the *Classroom Geographer;* this publicity has resulted in nearly 80 completed questionnaires. The respondents include 22 from water authorities and 31 from local weather enthusiasts, covering a wide range of geographical locations. Two issues of ROWS have already been produced, in September 1975 and January 1977.

Over the next five years it is hoped that the Register of Weather Stations project will collect information relating to most non-Meteorological Office weather stations. While funds are severely limited, growth will inevitably be slow but the organizers feel justified in pursuing what is regarded as a worthwhile goal.

We should like to point out that in the view of the Meteorological Office the process of making corrections to observations from non-standard sites is far from easy and may even be, in the strict sense, impossible. The relationships involved are non-linear and depend on all sorts of things such as wind direction, season etc.; for example, no wholly satisfactory relationship has yet been established between the temperature readings from the North Wall screen at

Kew and those from the standard screen. However, it is probably true to say that carefully made corrections to readings from a non-standard site will make them nearer than they otherwise would have been to those from a hypothetical standard site located in the same neighbourhood.

<div align="right">EDITOR</div>

REVIEW

Drought and Agriculture (*Report of the CAgM Working Group on the Assessment of Drought*), *WMO Technical Note*, No. 138, prepared by C. E. Hounam (Chairman), J. J. Burgos, M. S. Kalik, W. C. Palmer and J. [C.] Rodda. 275 mm × 210 mm, pp. xv + 127, *illus.* Secretariat of the World Meteorological Organization, Geneva, Switzerland. Price: 30 Sw. Fr.

This *Technical Note* is concerned with the complex subject of the effect of drought on agriculture, and on the whole succeeds well in highlighting the problems which must be considered in any analysis of agricultural drought. In particular it points out the distinction which must be drawn between 'aridity' and 'drought', since the agriculture of an area will have developed in such a way as to cope with the 'normal' situation, even if this means that there are long periods in each year when little or no rainfall occurs. Attention is drawn to the fact that rainfall is not the only determining factor in the timing and severity of a drought, but that evapotranspiration and soil moisture storage must also be taken into account, particularly in areas, such as the British Isles, where there is a large variation between winter and summer values of potential evaporation. The effects of drought depend also on the crop concerned, on its rooting characteristics, length of growing season, response to soil moisture stress and high levels of evaporative demand, and crop management.

The first three chapters contain a number of basic definitions and also examples of indices of agricultural drought. More complete details of these indices are given in Appendix I. Chapter 4 is concerned with methods of analysis of climatological data for drought studies. These chapters would provide the non-specialist in agricultural meteorology with a useful introduction to the subject, and make him aware of the still unsolved problems.

Chapters 5–9 deal with topics which are likely to be studied mainly by those primarily concerned with agricultural production; they include consideration of plant adaptation to drought conditions, water requirements of agriculture (including irrigation), diseases and pests in a drought, and local environmental control.

There are extensive bibliographies at the end of each chapter. Appendix II contains recommendations for a World Climate Watch on drought under the headings of historical assessment, real-time assessment, and prediction. Results could probably be achieved quite quickly for the first two for those areas where an adequate quantitative data base exists. However, a major breakthrough in the production of reliable forecasts of weather conditions over periods of from 5 to 30 days ahead is required if prediction of the continuation or cessation of a drought is to be achieved.

A drought which seriously affects agricultural production is likely to give rise to major economic, social and political consequences. Although this *Technical Note* has been written primarily for agricultural meteorologists and climatologists who may have to assess the drought risk for particular crops in particular areas, the introductory chapter (and to a lesser extent the two which follow it) could be read with profit by anybody who is concerned with problems of food production and distribution on a local or global scale. The Note is generally clearly and logically written, and provides a useful survey of the present state of knowledge on the subject of drought and agriculture, but in some parts too much minor detail of computation procedures is included, whilst elsewhere not enough information on basic assumptions is given.

MARJORY G. ROY

NOTES AND NEWS

Snow Survey of Great Britain

Survey Reports up to and including that for the 1967/68 season were published in *British Rainfall*. In future, Reports will be produced independently, normally in December each year. The Report for 1976/77 is now available from Meteorological Office Met O 3(b), London Road, Bracknell, Berks. RG12 2SZ at a price of £2 (post free) or a three-year advance subscription is offered at £5. Limited numbers of copies of the Reports for 1968/69 to 1975/76 are also available on application at a price of £1 per copy.

CONTENTS

NOTICES